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LETTER

On the surface impact of Arctic stratospheric ozone extremes

N Calvo[1], L M Polvani[2] and S Solomon[3]

[1] Departamento de Fisica de la Tierra II, Facultad de Ciencias Fisicas, Universidad Complutense de Madrid, Madrid, Spain
[2] Department of Applied Physics and Applied Mathematics, Department of Earth and Environmental Sciences, and Lamont Doherty Earth Observatory, Columbia University, New York, NY, USA
[3] Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA, USA

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Abstract

A comprehensive stratosphere-resolving atmospheric model, with interactive stratospheric ozone chemistry, coupled to ocean, sea ice and land components is used to explore the tropospheric and surface impacts of large springtime ozone anomalies in the Arctic stratosphere. Coupling between the Antarctic ozone hole and Southern Hemisphere climate has been identified in numerous studies, but connections of Arctic ozone loss to surface climate have been more difficult to elucidate. Analyzing an ensemble of historical integrations with all known natural and anthropogenic forcings specified over the period 1955–2005, we find that extremely low stratospheric ozone changes are able to produce large and robust anomalies in tropospheric wind, temperature and precipitation in April and May over large portions of the Northern Hemisphere (most notably over the North Atlantic and Eurasia). Further, these ozone-induced surface anomalies are obtained only in the last two decades of the 20th century, when high concentrations of ozone depleting substances generate sufficiently strong stratospheric temperature anomalies to impact the surface climate. Our findings suggest that coupling between chemistry and dynamics is essential for a complete representation of surface climate variability and climate change not only in Antarctica but also in the Arctic.

1. Introduction

The profound and manifold climate impacts caused by the depletion of stratospheric ozone over the South Pole in the last decades of the 20th century have been widely reported (Thompson *et al* 2011, Previdi and Polvani 2014). A large literature, comprising both observational and modeling studies, suggests that the formation of the ozone hole has affected nearly every component of the Southern Hemisphere climate system, from tropospheric winds and clouds, to surface temperatures and precipitation, and the circulation and ventilation of the Southern Ocean. In the Northern Hemisphere, however, connections have been less evident (e.g., Thompson and Solomon 2005).

The observed multi-decadal trends in ozone are much smaller over the Arctic than over the Antarctic (WMO 2014). Extensive Antarctic ozone depletion has occurred in all austral springs (with the possible exception of 2002) since the mid-1980s (WMO 2014). In contrast, the interannual variability in ozone is much larger in the Northern than in the Southern Hemisphere, owing to the abundance of planetary scale waves that propagate into the Arctic stratosphere and perturb the circulation there, leading to frequent 'sudden warming' events (Charlton and Polvani 2007). However, in years when the Arctic stratosphere is relatively unperturbed, anomalously weak transport by the Brewer–Dobson circulation (BDC) together with the concurrent anomalously low temperatures do conspire to yield extremely low ozone values in the spring. Such 'extreme ozone' events occur every few years: the latest one in 2011—with extremely cold temperatures from December to March and exceptionally large springtime Arctic ozone losses (Manney *et al* 2011)—has been extensively studied (see Solomon *et al* 2014, and references therein).

In contrast to the upper stratosphere where carbon dioxide increases have dominated temperature trends of recent decades, ozone depleting substances (ODS) are the main driver for lower stratospheric cooling trends that have been observed in the Arctic over the

last couple of decades (Rieder *et al* 2014). Years with greater depletion can be expected to display greater cooling, and vice versa: ozone and temperature are coupled. Here we examine whether Arctic stratospheric ozone minima are able to impact surface conditions, given what has been learned about the pervasive impact on the ozone hole on the Southern Hemisphere surface climate. Three previous studies have addressed that question.

Cheung *et al* (2014), using the UK Met Office operational weather forecasting system, explored the possible surface impacts associated with the 2011 extreme event (the largest on record). After substituting the standard ozone climatology in the model with the observed 2011 ozone from the Earth Observing System Microwave Limb Sounder, they found no significant reduction in tropospheric forecast errors in the month of March. Karpechko *et al* (2014) also investigated the possible surface response to the extremely low ozone anomalies in the spring of 2011, but with transient experiments performed with the atmospheric circulation model ECHAM5 (which has a well resolved stratosphere). They reported significant impacts on tropospheric climate only when stratospheric ozone anomalies were specified together with sea surface temperature for that year, but concluded that stratospheric ozone changes alone did not appear to have an effect on surface conditions. Finally, Smith and Polvani (2014) performed long, time-slice integrations with the Community Atmosphere Model, version 3 (CAM3)—to generate a large signal to noise ratio—and contrasted pairs of runs with high and low values of synthetic stratospheric ozone. They also found no significant tropospheric response to ozone differences for extreme amplitudes within the observed range of variability, but reported a statistically significant surface response when the amplitude was made somewhat larger than the observed 1979–2011 range.

From this one might be tempted to conclude that stratospheric ozone anomalies are simply unable to affect surface conditions. However, we note that these studies all suffer from several unphysical features that preclude such a hasty conclusion. First, all of these studies prescribed *zonal-mean* ozone fields in their models: there is now evidence that this considerably weakens the tropospheric response (Crook *et al* 2008, Gillet *et al* 2009, Waugh *et al* 2009). Second, they all used *monthly-mean* ozone fields, with daily ozone obtained via simple linear interpolation: Neely *et al* (2014) have recently shown that specifying daily ozone yields a substantial difference in years with large ozone losses, which amplifies the tropospheric and surface response to large ozone anomalies. Thirdly, all of these studies treat ozone as a *fixed external* forcing, and thus ignore any chemical–dynamical coupling: the severing of the relation between ozone concentrations and the stratospheric circulation implies that a potentially important feedback mechanism is absent from all those studies.

Recall that year-to-year variations in the amount of stratospheric ozone result from both dynamical and chemical processes (e.g. Tegtmeier *et al* 2008). Ozone is transported into the polar region by the BDC, and lost via heterogeneous chemical reactions on polar stratospheric clouds (PSC), which require cold temperatures below about 192 K in order to effectively deplete ozone. Winters with an anomalously cold polar stratosphere—associated with a strong polar vortex and a weak BDC—are more prone to the formation of PSCs, which destroy more ozone. This anomalous chemical ozone loss adds to the low dynamical ozone values associated with the weaker BDC transport from the tropics. As ozone further diminishes, the polar stratosphere cools further, yielding an even stronger polar vortex and a weaker BDC: hence the positive chemical–dynamical feedback. Conversely, in winters with an anomalously warm polar stratosphere, PSCs formation are obviously inhibited, thus reducing the loss of ozone. As more ozone builds up in the polar region, the temperatures there increase further which, in turn, prevents more heterogeneous chemistry and destruction of ozone, again providing a positive coupling between chemistry and dynamics.

In this paper we avoid the limitations present in earlier studies by using stratosphere-resolving atmospheric model with interactive stratospheric ozone chemistry, coupled to dynamical ocean, sea ice and land components. The ozone concentrations in our model, therefore, are fully consistent with the atmospheric circulation at all times, without any degradation from spatial or temporal averaging. Furthermore, the ozone chemistry is allowed to both respond to and feedback on the atmospheric circulation. Using an ensemble of model integrations from 1955 to 2005 with all known historical forcings, we find that the increase in ODS in the latter decades of the 20th century results in surface signals of Arctic ozone extremes that are robust and statistically significant, affecting surface winds, temperature and precipitation over large portions of the Northern Hemisphere.

2. Methods

In this study, we analyze output from the Whole Atmosphere Community Climate Model, Version 4 (WACCM4), one of the atmospheric components of the Community Earth System Model (CESM1). WACCM4 is a chemistry-climate model, with 66 levels in the vertical, and the model lid at 140 km. The vertical resolution is about 1.25 km in the troposphere and lower stratosphere, and up to 1.75 in the upper stratosphere; the horizontal resolution is 1.9° in latitude and 2.5° longitude. WACCM4 is run coupled to interactive ocean, sea ice and land components. The climate simulated by WACCM4 has been evaluated

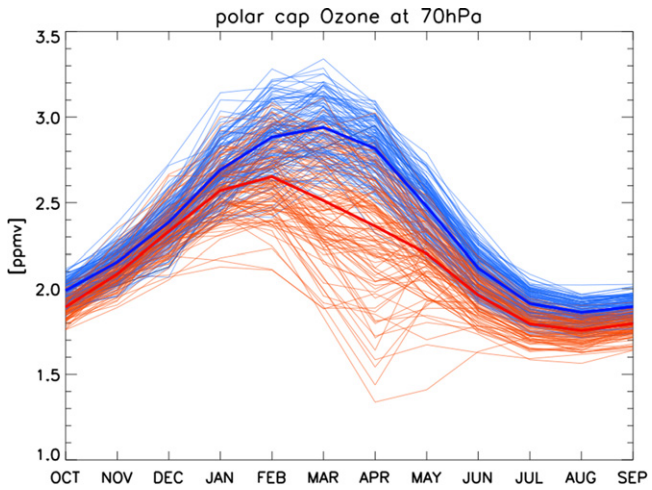

Figure 1. Seasonal evolution of polar cap ozone (averaged 65–90N) at 70 hPa. Thin lines: ozone from October of one year to September of the following. Blue: years from the period 1955 to 1975, using six simulations (for a total of 120 curves). Red: same as blue, but for the period 1985–2005. Thick lines are the means of the corresponding thin lines.

and documented in Marsh *et al* (2013), which the reader may consult for other details about the model setup.

A total of six historical WACCM4 simulations, from 1955 to 2005, similar to those analyzed in Marsh *et al* (2013), are analyzed here. These simulations incorporate all historical forcings as specified by the Climate Model Intercomparison Project, Phase 5 (CMIP5), which include observed greenhouse gases and halogen concentrations, total spectral irradiance, and volcanic aerosol. The six simulations only differ in their initial conditions, and were conducted with an imposed quasi-biennial oscillation, which greatly improves the representation of ozone variability in the tropical stratosphere. Note that using six 50-year long runs, we have 300 winters at our disposal: this is a much larger sample that is available in the observations, and it provides a large signal-to-noise ratio allowing us to extract surface signatures of large stratospheric ozone anomalies that may not be mirrored in the single realization available in observations; we return to this point below. We have checked that the members of the ensemble are statistically independent by computing the correlation between the daily time series of Arctic polar cap temperature at 50 hPa: the value of the correlation never exceeds 0.1, as one might expect from the large internal dynamical variability which largely controls the occurrence of extreme ozone events.

3. Results

We start by illustrating the large interannual variability in the seasonal cycle of Arctic stratospheric ozone. Each curve in figure 1 shows the simulated time evolution of polar cap ozone (averaged between 65 and 90N) at 70 hPa, from October of one year to September of the following year. The largest year-to-year differences occur in the spring, notably in April, when values range from 1.3 ppmv to 3.1 ppmv. As discussed above, these large interannual differences appear as a consequence of the coupling between dynamics and chemistry. As shown in supplementary figure S1, the range of WACCM4 ozone in Arctic compares favorably with ozonesonde observations.

Next we focus on the crucial role of ODS in producing extremely low-ozone years. We do this by considering separately the first and last two decades of our simulations, which we illustrate in figure 1 by color-coding years from 1955 to 1975 in blue, and years from 1985 to 2005 in red (the winters in the middle decade, 1975–1985, are not shown). The thick blue and red curves show the average of the corresponding 120 curves (20 years x 6 simulations). Notice that there is a very clear separation between the early and late decades. In the early period (blue curves), i.e. before the large increase in ODS, the interannual variability is relatively modest, and it is largely due to dynamics. In the late period (red curves) the year-to-year spread becomes much larger, as a consequence of the higher concentrations of ODS. Furthermore, notice that while high ozone winters occur in both periods, the very low ozone winters are seen exclusively in the last two decades, i.e. after 1985, as ODS are needed to produce extreme ozone depletion. This is consistent with observations (see, for instance, Dameris *et al* 2014), and the extreme low values of about 1.5 ppmv in April at 70 hPa agree with satellite data for 2011 reported in Manney *et al* (2011).

One might wonder how the ranges between high and low ozone years in our WACCM4 simulations compare with those explored in Smith and Polvani (2014; see their figure 1), who used highly idealized 'synthetic' ozone concentrations in order to control the amplitude of the high/low range (they focused on

±15% and ±25% of the March total column ozone). With reference to our own figure 1, the range in the early period—computed from the 70 hPa polar cap ozone difference in April between the 30 highest and the 30 lowest blue curves—is approximately ±10%, whereas for the late period the range is approximately ±20%. These numbers are consistent with the findings Smith and Polvani (2014), although one ought to keep in mind that the ozone used in that study was specified from zonally symmetric and monthly mean values and, furthermore, that the ozone was decoupled from the dynamics in that study.

To examine whether the dynamics is fundamentally different between the earlier and later periods, we have computed the statistics of stratospheric sudden warmings (SSW) events for the two periods, following the simple method of Charlton and Polvani (2007). We find that the frequency of occurrence of SSWs, approximately 4.5 events per decade in WACCM (as already reported in Marsh *et al* 2013), is identical—in terms of statistical significance—between the 1955−1975 and the 1985–2005 periods. Hence the presence of elevated levels of ODS, and the accompanying ozone depletion, do not seem to affect the frequency of SSWs.

We now turn to the main question of interest, namely whether one can detect a signature of the extreme ozone anomalies in the troposphere and at the surface. We do this following the methodology of Smith and Polvani (2014), which consists of taking the difference between the extremes of high and low ozone years. To bring out the key role of ODS, we do this separately for the early (1955–1975) and the late (1985–2005) periods. Since we have 120 years available for each period, we bring out the ozone signal by simply subtracting in each period the 30 highest from the 30 lowest winters at 70 hPa in April, when the differences are the largest (see figure 1). The results are shown in figure 2, with the differences for the early period shown in the left column, and for the late period on the right.

The vertical structure of the polar cap ozone differences, as a function of month, is shown in the top row of figure 2: they are, of course, much stronger for the late period (panel b), as already shown, owing to the high levels of ODS. Accompanying the ozone differences, one finds clear polar cap temperature differences, shown in the middle row. Again, they are much stronger in the late period (panel d), as one might expect. More surprisingly, perhaps, is the fact that they are only barely significant in the early period (panel c; significance is indicated by stippling). This is because, in the absence of ozone depletion, the interannual ozone variability is not sufficiently large to produce a clear temperature difference between the high and low ozone years (note the relatively small spread of the blue curves in figure 1).

The novel result of our study becomes apparent once we move from temperature to wind differences, shown in the bottom row of figure 2. For the late period (panel f), the ozone and temperature signals (which peak in March and April) are accompanied by a very strong signal in the winds (here averaged 60–75N), which extends all the way to the surface in April and May. This agrees well with the findings of Smith and Polvani (2014), who also found that the large signals due to stratospheric ozone differences appear below the tropopause in April and May. Recall, however, that the ozone values had to be artificially inflated in that study and the model they used lacked a well-resolved stratosphere and interactive stratospheric chemistry. In contrast, for the simulations analyzed here, all forcings are consistent with observations and have realistic values: the large ozone anomalies that produce a clear surface signal in WACCM4 are a natural consequence of the positive feedback between chemistry and dynamics. Needless to say, in the early period, when ODS concentrations are low, no statistically significant wind signal is found, even in the stratosphere (panel e).

It is interesting to document the role of SSWs in producing the differences between high- and low-ozone composites in figure 2. Although, as mentioned above, the frequency of SSWs in our model (about 4.5 events per decade) is the same in the early and late periods, the role of wintertime SSWs in determining the spring ozone anomalies appears to be fundamentally different between the two periods. In the early period, we find little difference in SSW frequency between the 30 high-ozone and the 30 low-ozone years (12/30 for the high-ozone years, and 10/30 for the low-ozone years have one or more SSWs in that winter). In contrast, for the later period, the frequency of SSWs is very different between the high and low ozone years (19/30 and 0/30, respectively). This clearly show that, in the presence of elevated ODS concentrations, the occurrence of SSWs becomes crucial for determining the April ozone anomalies, whereas SSWs play a relatively minor role in the early period.

The zonally-averaged picture presented in figure 2 is severely limiting, as the tropospheric and surface signature of ozone anomalies possesses considerable regional structure. To illustrate this, focusing now on April–May when the signal is largest at the surface, we now turn to latitude–longitude plots, which reveal the full complexity of the tropospheric and surface signatures. In figure 3 (top row) we show the differences in zonal mean zonal wind at 500 hPa, between the low and high ozone years. For the early period (panel a) there is little to see, but for the late period (panel b) a very robust signal can be seen over the North Atlantic and Eurasia, corresponding to a poleward shifted jet, as one might expect to accompany a strong vortex (which is typically associated with a low-ozone year).

Going from the mid-troposphere to the surface, one can see the extreme ozone signature in the April–May differences in sea level pressure (SLP, figure 3, middle row). The structure is reminiscent of a positive

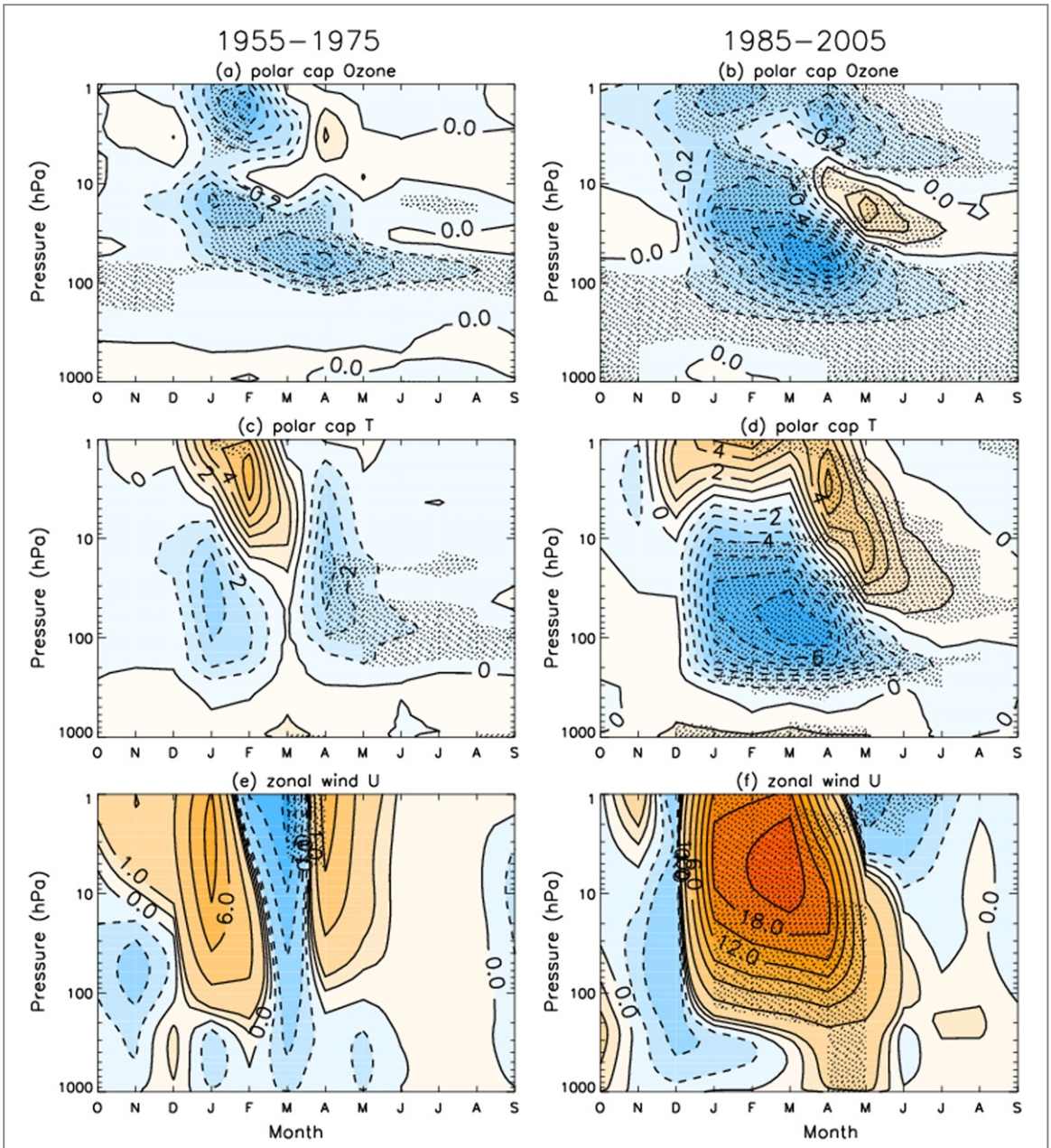

Figure 2. Time-height evolution of the composite differences (LOW-HIGH ozone) between the 30 years with the lowest polar cap ozone values and the 30 years with the highest polar cap ozone values at 70 hPa in April (see figure 1). (a), (b) Polar cap (65–90N) ozone. (c), (d) Polar cap (65–90N) temperature. (e), (f) 65–75N zonal mean zonal wind. Left: composite differences for the 1955–1975 period; right: for the 1985–2005 period. Units are ppmv for ozone, K for temperature and m s^{-1} for wind. Stippling indicates significant differences at the 95% confidence level (using a Student t-test).

Northern Annual Mode (NAM) pattern, in agreement with the strengthening of the zonal winds coming from the stratosphere. Negative SLP values appear in the polar cap region and positive values at middle latitudes, mainly in the Atlantic and the Pacific coast of Asia. Thus, two dipoles are clearly simulated during April–May, one in the Atlantic suggestive of the positive phase of the North Atlantic Oscillation, and the other in eastern Asia. We note, however, the absence of positive anomalies over the North Pacific, which are typical in a positive NAM phase. Together with the positive anomaly over eastern Asia, the ozone signal in SLP is closer to the pattern of the seasonal varying NAM, which is somewhat different from the typical winter NAM (e.g. Ogi *et al* 2004).

The April–May surface temperature differences arising from stratospheric ozone anomalies are shown in the bottom row: again, large significant areas are only found for the 1985−2005 period (right column, panel f). Two robust features stand out. First, a significant warming in Northern Europe and Siberia, with the largest values over this region in excess of 5 K; note also the accompanying cooling that appears over China, consistent with a positive NAM phase. Second, we note significant cooling in Greenland and the northeastern part of Canada, again with an amplitude

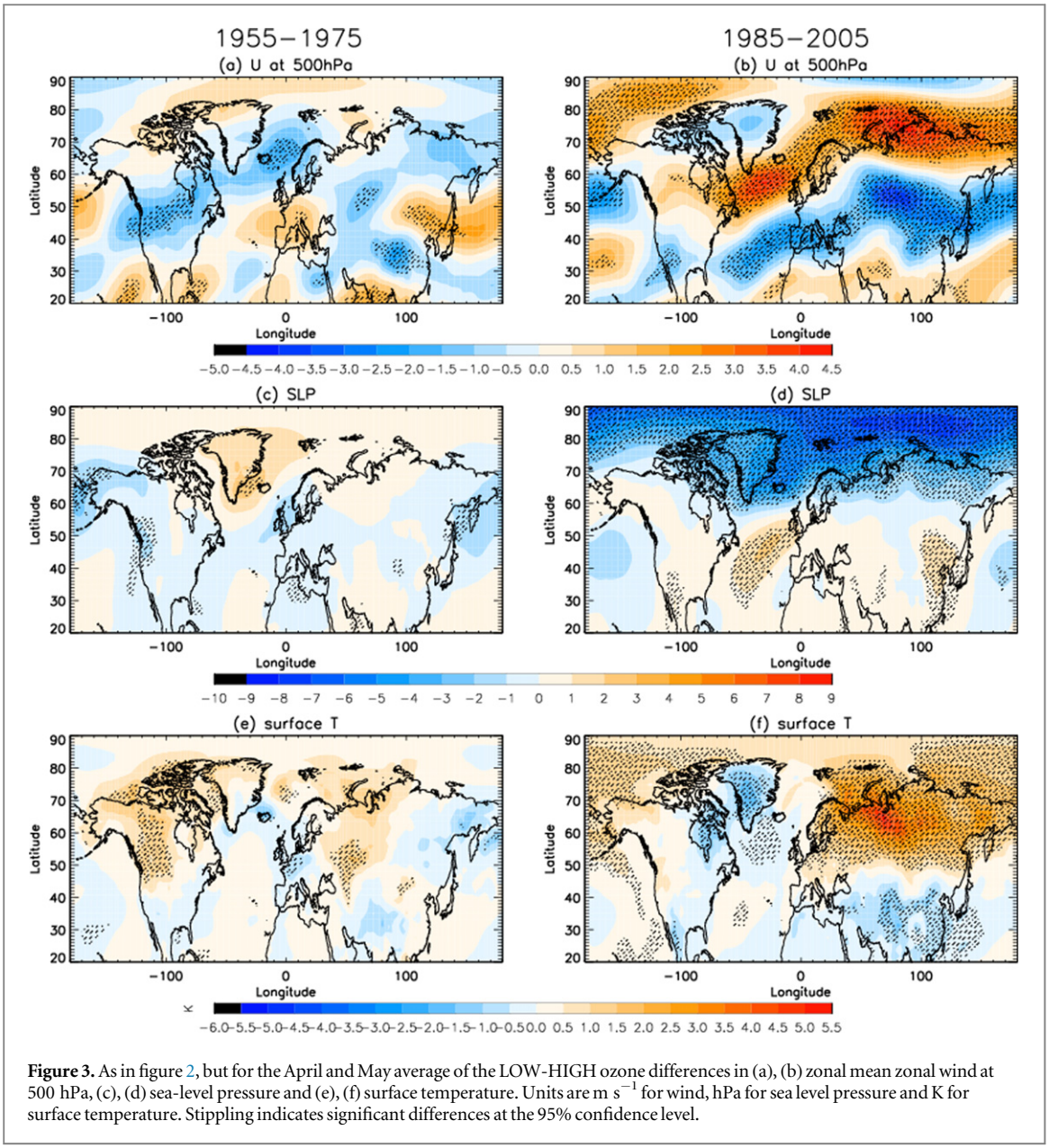

Figure 3. As in figure 2, but for the April and May average of the LOW-HIGH ozone differences in (a), (b) zonal mean zonal wind at 500 hPa, (c), (d) sea-level pressure and (e), (f) surface temperature. Units are m s^{-1} for wind, hPa for sea level pressure and K for surface temperature. Stippling indicates significant differences at the 95% confidence level.

of several degrees. Interestingly, both of these features are identical (and of opposite sign, as one might expect), to those observed on a seasonal basis following SSW events (see, e.g., figure 2 of Butler *et al* 2014), when they are associated with an anomalously weak polar vortex and anomalously high levels of stratospheric ozone.

Finally, recalling that changes in the location of the jet stream are usually accompanied by changes in precipitation, we conclude by showing differences in the latter that arise from the ozone anomalies. Since the tropospheric signal of such anomalies is far from zonally symmetric, as we have just shown, a signal in the zonal mean precipitation field is not expected. Keeping in mind the patterns in figure 3(b), we focus on two regions: the Atlantic and Eurasia, where the ozone signals appear at distinct latitudes. In figure 4 we plot the April–May difference, separately, over these two areas (70W to 20E for the North Atlantic Ocean, panel a; and 50E to 120E for the Siberian region, panel b). In both regions, a significant enhancement in precipitation at high latitudes can be seen for the period 1985–2005 (red curves), associated with extremely low polar stratospheric ozone. This response is largely absent in the earlier period (blue curves), as the ozone anomalies prior to the large anthropogenic ODS increases are not sufficiently strong to be significant at the surface.

4. Conclusion

We have shown that the year-to-year variability of lower stratospheric ozone over the Arctic, which arises when the internal variability of the stratosphere-troposphere system occurs in the presence of high levels of anthropogenic ODS, can have a significant impact on surface climate. Specifically, extreme ozone

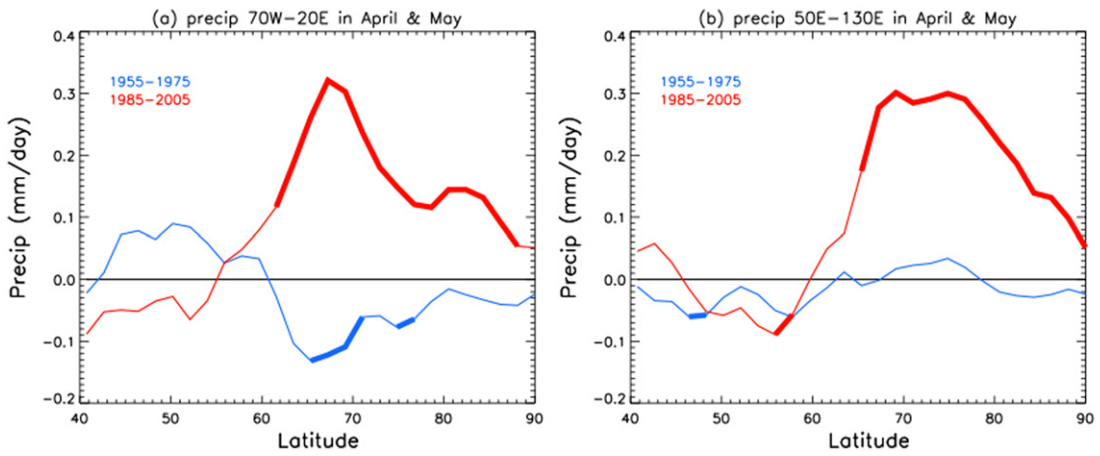

Figure 4. Simulated precipitation difference (LOW-HIGH ozone) averaged for April and May (in mm day^{-1}) over the region (a) between 70W and 20E and (b) 50E and 130E. Blue lines show model results for the 1955–1975 period, while red lines are for the 1985–2005 period. Thick curves indicate significant differences at the 95% confidence level.

loss events over the North Pole—as simulated in a ensemble of simulations by a state-of-the-art chemistry-climate model with standard CMIP5 forcings typical of the late decades of the 20th century—are able to affect surface pressure, temperature and precipitation over much of the Northern Hemisphere. Ensemble simulations are key to our findings, allowing sufficient sampling to clearly identify the effects of ozone changes despite the many other factors that lead to high interannual variability in tropospheric climate. Thus the use of such model simulations provides a basis for analysis that is a far more sensitive test than the real world's single realization. Compounding what has been learned over the Southern Hemisphere in the last decade, our results further highlight the importance of stratospheric processes for a complete understanding of surface climate variability and climate change in both hemispheres.

Our study also shows that when the levels of ODS are relatively low (as in the period 1955–1975) the dynamical variability alone is unable to produce ozone extremes of sufficient magnitude to be significant at the surface. This suggests that the coupling between chemistry and dynamics, therefore, might be essential for the amplification of the stratospheric signal and its propagation into the troposphere. The importance of chemistry–dynamics coupling has been noted in prior studies (e.g. Milewski and Bourqui 2011). In order to properly quantify the magnitude of this coupling, a direct comparison of coupled and uncoupled model integrations is needed. We do not have those at our disposal, but plan to explore this issue in future papers.

Our results have clear implications for weather forecasting. The recent study of Cheung *et al* (2014), which focused on March of 2011 when the largest Arctic ozone minimum in the record has been observed, concluded that merely specifying observed (and very low) ozone concentrations is not sufficient to see any reduction in forecast errors in the UK Met Office operational forecast model, although that study did not present results for later times in the spring. Our results suggest that the absence of coupling between chemistry and dynamics in that model might explain the negative result. Whether the inclusion of stratospheric chemistry, perhaps in some simplified form with a reduced number of species, is practically feasible in any existing operational forecast model remains to be examined. The findings of our study offer clear evidence that this possibility should be explored.

Acknowledgments

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